Dynamical Climatology

Transport mechanisms in the middle atmosphere: an introductory survey.

by

D.G. Andrews

DCTN 51

February 1987

Meteorological Office (Met. O. 20)
London Road
Bracknell
Berkshire RG12 2SZ

DYNAMICAL CLIMATOLOGY TECHNICAL NOTE NO. 51

TRANSPORT MECHANISMS IN THE MIDDLE ATMOSPHERE: AN INTRODUCTORY SURVEY

by

David G Andrews

Submitted for publication in Advanced Studies Institute Series

Met O 20 (Dynamical
Climatology Branch)
Meteorological Office
London Road
Bracknell
Berkshire, RG12 2SZ U.K.

February 1987

NOTE: This paper has not been published. Permission to quote from it should be obtained from the Assistant Director of the above Meteorological Office branch.

Transport mechanisms in the middle atmosphere: an introductory survey

David G. Andrews
Meteorological Office Unit, Hooke Institute
Clarendon Laboratory, Parks Road,
Oxford OX1 3PU

A paper presented at the NATO Advanced Study Institute
'Transport Processes in the Middle Atmosphere',
Erice, Sicily, 23-27 November 1986.

TRANSPORT MECHANISMS IN THE MIDDLE ATMOSPHERE : AN INTRODUCTORY SURVEY

David G. Andrews
Meteorological Office Unit
Hooke Institute
Clarendon Laboratory
Parks Road
Oxford OX1 3PU
United Kingdom

ABSTRACT. Basic ideas concerning the transport of dynamical and chemical tracers in the middle atmosphere are presented. Tracer transport models in one, two and three dimensions are discussed and some of the conceptual and practical advantages and disadvantages of each type are indicated.

1. INTRODUCTION

In this section we present some basic definitions pertaining to atmospheric transport and give some examples of transported quantities. The most fundamental concept is that of a materially conserved tracer (often called a conservative tracer), the amount of which remains constant in time following each moving mass element of air (or 'air parcel'). If χ is the 'amount' of tracer per unit mass of air then the material conservation property has the mathematical expression

$$\frac{D\chi}{Dt} \equiv \frac{\partial\chi}{\partial t} + \underset{\sim}{u}\cdot\nabla\chi = 0, \qquad (1.1)$$

where $\underset{\sim}{u}$ is the fluid velocity and t is time. In practice, no physical tracer is precisely conserved and (1.1) must be replaced by

$$\frac{D\chi}{Dt} = S, \qquad (1.2)$$

where S represents sources and sinks of tracer. However, for many atmospheric tracers S is weak in the sense that it only leads to small relative changes in χ following an air parcel, over relevant advection timescales: χ is then called a quasi-conservative or long-lived tracer.

A second way of classifying tracers is to distinguish between dynamically-active tracers, whose dynamical, thermodynamic or chemical properties allow them to influence the flow field that transports them, and dynamically-passive tracers, which have no such effect. In principle

the latter could be chemically active.

The most well-known atmospheric tracer is the potential temperature θ, which is related to temperature T and pressure p by the relation

$$\theta \equiv T(p/p_s)^{-\kappa} , \tag{1.3}$$

where p_s is a reference pressure (e.g. 1000 mb) and $\kappa = R/c_p \simeq 2/7$, R being the gas constant for dry air and c_p the specific heat at constant pressure. (The entropy per unit mass is then $c_p \ell n\theta$ + constant.) This satisfies the equation

$$\frac{D\theta}{Dt} = \frac{J}{c_p}\left(\frac{p}{p_s}\right)^{-\kappa} \equiv Q, \tag{1.4}$$

where J is the net heating rate per unit mass. θ is quasi-conservative for sufficiently weak heating rates -- for example if J represents net radiative heating in a region where the radiative relaxation time is long compared to the advection time. Since θ is a basic thermodynamic variable, it is clearly a dynamically-active tracer.

Somewhat less familiar is Ertel's potential vorticity P, defined by

$$P \equiv \rho^{-1}(\underset{\sim}{\omega} + 2\underset{\sim}{\Omega})\cdot\nabla\theta , \tag{1.5}$$

where ρ is atmospheric density, $\underset{\sim}{\omega} \equiv \nabla\times\underset{\sim}{u}$ is the relative vorticity and $\underset{\sim}{\Omega}$ is the Earth's rotation vector. If a body force $\underset{\sim}{X}$ per unit mass acts on the atmosphere then

$$\frac{DP}{Dt} = \rho^{-1}(\underset{\sim}{\omega} + 2\underset{\sim}{\Omega})\cdot\nabla Q + \rho^{-1}\nabla\theta\cdot\nabla\times\underset{\sim}{X} , \tag{1.6}$$

where Q is defined in (1.4). P is quasi-conservative if the heating term Q and the body force $\underset{\sim}{X}$ are sufficiently weak. P is also a dynamically-active tracer in general: this follows from the 'invertibility principle', which states that the field of P at any instant determines the velocity and temperature fields at that instant, given suitable boundary conditions and the assumption that the atmosphere is in a 'balanced' state. [For discussion of this principle and many other aspects of the use of Ertel's potential vorticity, see the review by Hoskins, McIntyre and Robertson (1985).]

A further broad class of atmospheric tracers comprises the minor chemical species, with χ given by the mass or volume mixing ratio. Many of these species are long-lived: examples include ozone in the lower stratosphere and methane and nitrous oxide throughout most of the stratosphere. Much of the current interest in tracer transport is stimulated by the need for better understanding of the ways in which these chemicals are carried from one part of the atmosphere to another, and especially how they move in the latitudinal and vertical directions from source regions to sinks and reservoirs. Other papers in this volume are devoted to specific examples of these processes.

2. SOME BASIC TRANSPORT MECHANISMS

To illustrate some of the ways in which passive, conservative or quasi-conservative tracers can be transported by atmospheric motions, we now consider some very idealised flows. We first restrict attention to two-dimensional flows $u = (u(x,y), v(x,y))$ in the horizontal plane, where x is distance eastwards and y is distance northwards from some origin. In the simplest case in which u = constant and v = 0, the flow is a uniform eastward advection and a blob of passive, conservative tracer released into the fluid will be carried bodily eastward, with no distortion from its initial shape. If however u varies with y (but v = 0 still), the blob will become stretched out in longitude by the shear, some parts being transported more rapidly than others. The longitudinal extent of the blob grows linearly with time and sharp y-gradients of tracer mixing ratio develop; however, no latitudinal transport can take place.

Suppose now that the tracer is weakly diffusive, rather than precisely conservative, with the source term S in (1.2) given by

$$S = \lambda \nabla^2 \chi \ ,$$

where λ is a small, constant diffusion coefficient. For a blob released into the zonal flow $u = (u(y), 0)$, the combination of the y-shear and the latitudinal diffusion tends to lead to a rapid effective zonal diffusion (Taylor, 1953); hence the tracer tends to homogenise in the x-direction. Transport in the y-direction is still weak, however.

A further useful idealisation is to allow for a slow latitudinal flow v = constant, in addition to the zonal flow u(y) and diffusion. If v is small enough, the longitudinal homogenisation occurs before v has had much effect. Eventually, however, v does lead to significant meridional transport of the zonally-symmetrized tracer field.

Flows that are not zonally-symmetric are much more effective at bringing about latitudinal transport. Consider for example the steady flow depicted in Fig.1, which includes a meandering 'jet stream'. If tracer is injected at a low-latitude source region and removed in a high-latitude sink, a systematic poleward transport is performed by this wavy flow in conjunction with non-conservative processes (Pyle and Rogers, 1980).

Of course, the real middle atmosphere -- especially the northern winter stratosphere -- is much more complex still, with rapid meridional transport by transient, large-amplitude planetary-wave motions. This is vividly illustrated by maps of quasi-conservative tracers derived from satellite measurements: see for example McIntyre and Palmer (1983, 1984) and Clough, Grahame and O'Neill (1985) for maps of Ertel's potential vorticity on isentropic surfaces, Jones (1984) for methane maps and Leovy et al.(1985) for ozone maps. (In particular, the latter paper indicates how ozone is transported polewards, from photochemical source regions in low latitudes to high northern latitudes, in narrow tongues associated with the transient major and minor warming events occurring in the northern winter.) Much current work is being devoted to attempts to quantify the information provided by such maps, and interpret them in

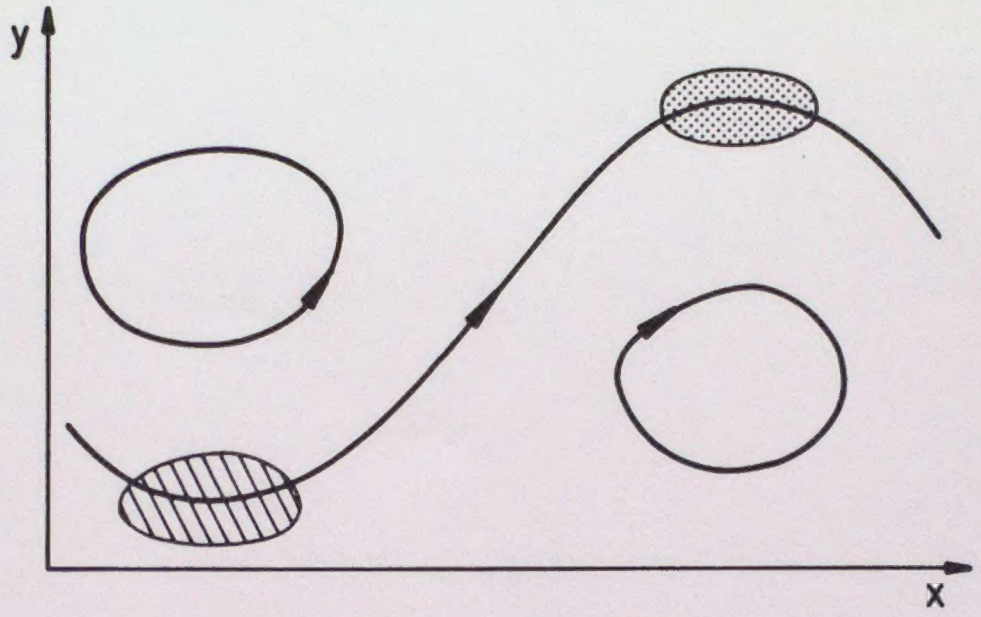

Figure 1. Illustrating how a steady meandering jet stream can transport tracer meridionally from a low-latitude source (shown hatched) to a high-latitude sink (shown stippled).

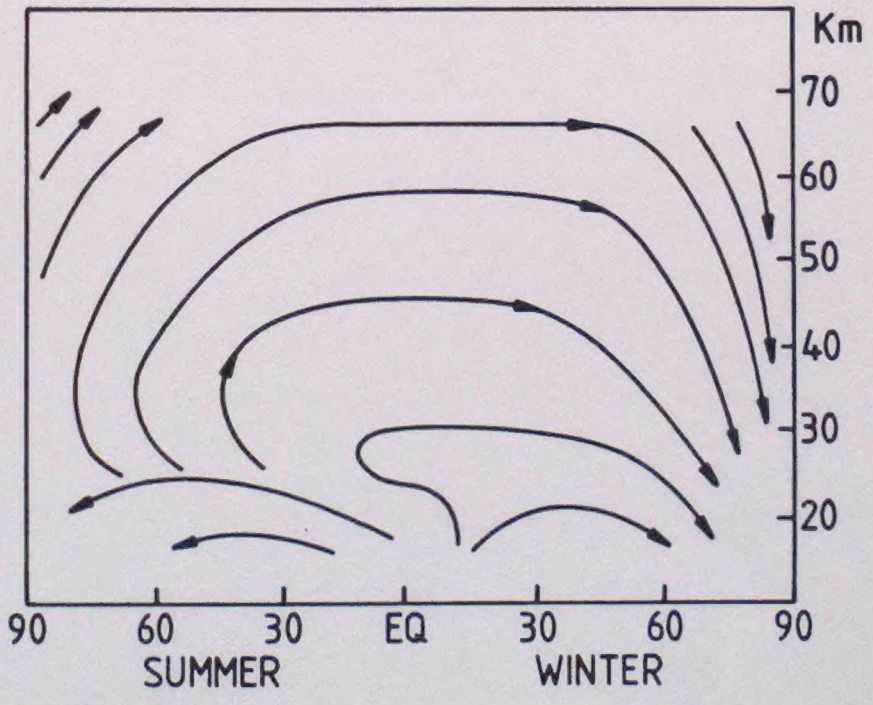

Figure 2. Schematic representation of Murgatroyd and Singleton's diabatic circulation at the solstices. (After Dunkerton, 1978.)

dynamically-meaningful ways (e.g. Butchart and Remsberg, 1986).

In addition to the quasi-horizontal, large-scale transport depicted by these maps, significant vertical transport of tracers may be brought about by other types of wave, notably breaking gravity waves in the upper mesosphere: see, e.g., Garcia and Solomon (1985).

3. MODELS OF TRANSPORT

3.1 One-dimensional models

One-dimensional models have been much used by atmospheric chemists to predict the vertical structure of constituents. To demonstrate this approach we introduce the log-pressure vertical coordinate

$$z \equiv -H\ell n(p/p_s), \tag{3.1}$$

where H is a constant, representative scale height (typically about 7 km in the middle atmosphere). Then the mass-continuity equation in Cartesian coordinates is

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \rho_0^{-1}\frac{\partial(\rho_0 w)}{\partial z} = 0, \tag{3.2}$$

where $w = Dz/Dt$ and $\rho_0(z) = p/gH \propto e^{-z/H}$, g being the acceleration of gravity. Using (3.2) the tracer continuity equation (1.2) can be written in 'flux form'

$$\frac{\partial \chi}{\partial t} + \frac{\partial}{\partial x}(\chi u) + \frac{\partial}{\partial y}(\chi v) + \rho_0^{-1}\frac{\partial}{\partial z}(\rho_0 \chi w) = S \ . \tag{3.3}$$

The horizontal global integral of any quantity ψ is

$$\langle\psi\rangle \equiv \iint_{\text{globe}} \psi(x,y,z,t)\, dx\, dy \ ; \tag{3.4}$$

$\langle\psi\rangle$ is a function of z and t. On applying (3.4) to (3.3) and noting that the global integral of the horizontal divergence vanishes, we obtain

$$\frac{\partial\langle\chi\rangle}{\partial t} = -\rho_0^{-1}\frac{\partial}{\partial z}(\rho_0\langle\chi w\rangle) + \langle S\rangle \ . \tag{3.5}$$

The term $\langle\chi w\rangle$ represents the globally-integrated vertical flux of tracer. To obtain a prognostic equation for $\langle\chi\rangle$ it is usually assumed that this flux can be represented by diffusion down the (vertical) gradient of $\langle\chi\rangle$; thus

$$\langle\chi w\rangle = -K(z)\frac{\partial\langle\chi\rangle}{\partial z} \ , \tag{3.6}$$

where K(z) is an empirical diffusion coefficient. On substituting (3.6) into (3.5) we obtain a diffusion equation for $\langle\chi\rangle$, with sources and sinks:

$$\frac{\partial\langle\chi\rangle}{\partial t} = \rho_0^{-1}\frac{\partial}{\partial z}\{\rho_0 K\frac{\partial\langle\chi\rangle}{\partial z}\} + \langle S\rangle \ . \qquad (3.7)$$

This approach can be generalised to allow for many constituents χ_n (n = 1,2,...N) where the relevant sources $\langle S_n\rangle$ may depend on all the $\langle\chi_m\rangle$; in general a different K_n should be used for each constituent.

The advantage of the one-dimensional approach is that the representation of transport is so simple that the available computer power can be devoted mainly to consideration of a large number of chemical species and many reactions. Models of this kind are therefore valuable for testing and comparing photochemical theories. However, since such models deal with global (or sometimes hemispheric) integrals, it is difficult to justify their comparison with local measurements of constituents in the real atmosphere. It is also difficult to give a convincing justification for the 'diffusive flux' parameterisation (3.6), although recent papers by Holton (1986) and Mahlman, Levy and Moxim (1986) have provided some rationale for such an approach.

3.2 Two-dimensional models

A less restrictive approach is to employ the zonal (Eulerian) average $\overline{\Psi}(y,z,t)$. Although this is again a nonlocal quantity, the zonal-homogenisation process mentioned above may frequently be relevant to the middle atmosphere, in the absence of strong planetary-wave motions. Moreover, satellite measurements of composition are often averaged zonally to improve signal-to-noise levels. A study of the transport of $\overline{\chi}$ in the meridional plane is therefore of some interest.

We denote deviations from the zonal mean (loosely referred to as the 'eddy' or 'wave' contributions) by primes; (3.2) and (3.3) then yield the 'Eulerian mean' equation

$$\frac{\partial\overline{\chi}}{\partial t} + \overline{v}\frac{\partial\overline{\chi}}{\partial y} + \overline{w}\frac{\partial\overline{\chi}}{\partial z} = -\left[\frac{\partial}{\partial y}(\overline{v'\chi'}) + \rho_0^{-1}\frac{\partial}{\partial z}(\rho_0\overline{w'\chi'})\right] + \overline{S} \qquad (3.8)$$

after zonal averaging. This equation shows that the rate of change of $\overline{\chi}$ is determined not only by the zonal-mean sources and sinks and the advection of $\overline{\chi}$ by the Eulerian-mean wind $(\overline{v}, \overline{w})$, but also by an 'eddy flux divergence' (in square brackets). In practice, it is often found that the 'mean advection' and 'eddy flux' terms almost cancel: the reason is that the Eulerian mean circulation $(\overline{v}, \overline{w})$ is not independent of the presence of the eddies -- indeed it is usually largely determined by them. One should therefore not attach much physical significance to the labels 'mean transport' and 'eddy transport' that are often given to these expressions. In fact it turns out that there are other, quite different, ways of arranging eq.(3.8) into 'mean advective transport' and 'eddy flux divergence' terms: these will be discussed below. Before doing this, however, we recall some historical details.

Some early ideas about meridional transport of tracers in the lower stratosphere were introduced by Brewer (1949) and Dobson (1956), who suggested that observed distributions of water vapour and ozone might qualitatively be accounted for by advection by a hypothetical zonally-symmetric circulation. This flow consists of a rising branch passing through the tropical tropopause, and poleward and downward flow in the extratropics; it has come to be known as a 'Brewer-Dobson circulation'.

A few years later, Murgatroyd and Singleton (1961) performed a diagnostic calculation of a 'possible meridional circulation' in the middle atmosphere, here denoted by $(\overline{v}_D, \overline{w}_D)$, from the spherical-geometry equivalents of the zonal-mean thermodynamic equation

$$\frac{\partial \overline{T}}{\partial t} + \overline{v}_D \frac{\partial \overline{T}}{\partial y} + \overline{w}_D \left(\frac{\partial \overline{T}}{\partial z} + \frac{\kappa \overline{T}}{H} \right) = \frac{\overline{J}}{c_p} \qquad (3.9)$$

and mass-continuity equation

$$\frac{\partial \overline{v}_D}{\partial y} + \rho_0^{-1} \frac{\partial}{\partial z} (\rho_0 \overline{w}_D) = 0, \qquad (3.10)$$

together with boundary conditions $\overline{v}_D = 0$ at the poles. They obtained the requisite $\overline{T}(y,z,t)$ from observations and the net heating rate $\overline{J}$ from a radiative calculation using $\overline{T}$ and observed ozone fields. However, they did not include any eddy heat flux divergence terms in the thermodynamic equation (3.9). A sketch of the resulting circulation at the solstices is given in Fig.2: in the lower stratosphere it resembles the Brewer-Dobson circulation, while higher up it consists of rising motion in the summer hemisphere, cross-equatorial flow in the mesosphere and descent in the winter hemisphere. A circulation diagnosed by this method is now often called a 'diabatic circulation'. (The $\partial \overline{T}/\partial t$ term is usually small, and is sometimes omitted from the calculation.)

Murgatroyd and Singleton noted that their results, being based on zonal-mean observations, necessarily omitted the 'eddy heat flux convergence' term

$$C \equiv - \frac{\partial}{\partial y} (\overline{v'T'}) - \rho_0^{-1} \frac{\partial}{\partial z} (\rho_0 \overline{w'T'}) - \frac{\kappa \overline{w'T'}}{H} \qquad (3.11)$$

that would appear on the right of (3.9) if $(\overline{v}_D, \overline{w}_D)$ were to be replaced by $(\overline{v}, \overline{w})$. Thus their diabatic circulation differs from the Eulerian-mean circulation in regions where C is significant, such as the northern winter stratosphere: in this region Eulerian-mean ascent, rather than descent, is found.

We now return to the alternative formulations of the zonal-mean tracer equation (3.8) mentioned above. One of these is the so-called 'transformed Eulerian-mean' approach, in which the 'residual circulation' $(\overline{v}^*, \overline{w}^*)$, defined by

$$\overline{v}^* \equiv \overline{v} - \rho_0^{-1} \frac{\partial}{\partial z} (\rho_0 \Psi), \quad \overline{w}^* \equiv \overline{w} + \frac{\partial \Psi}{\partial y}, \qquad (3.12)$$

is introduced, where

$$\Psi \equiv \overline{v'T'} \Big/ \left(\frac{\partial \overline{T}}{\partial z} + \frac{\kappa \overline{T}}{H}\right) \tag{3.13}$$

(Andrews and McIntyre, 1976, 1978a). Then in place of (3.8) we have

$$\frac{\partial \overline{\chi}}{\partial t} + \overline{v}^* \frac{\partial \overline{\chi}}{\partial y} + \overline{w}^* \frac{\partial \overline{\chi}}{\partial z} = E + \overline{S} \tag{3.14}$$

where $\rho_0 E$ is the divergence of a mean quadratic function of eddy quantities. The relevance of the residual circulation in the present context is that it satisfies not only a mass continuity equation identical in form to (3.10) but also a thermodynamic equation

$$\frac{\partial \overline{T}}{\partial t} + \overline{v}^* \frac{\partial \overline{T}}{\partial y} + \overline{w}^* \left(\frac{\partial \overline{T}}{\partial z} + \frac{\kappa \overline{T}}{H}\right) = G + \frac{\overline{J}}{c_p} \tag{3.15}$$

with a mean quadratic eddy term G. In contrast to the quantity C defined in (3.11), G is formally negligible for quasigeostrophic motions. In such a case, and indeed under a wider variety of circumstances as well, $(\overline{v}_D, \overline{w}_D) \approx (\overline{v}^*, \overline{w}^*)$; thus Murgatroyd and Singleton's diabatic circulation is a good approximation to the residual circulation in many cases of atmospheric interest. It is also quite closely related to the 'generalised Lagrangian-mean' circulation (Andrews and McIntyre, 1978b) defined, roughly speaking, as an average following sets of moving fluid parcels rather than along lines of constant latitude and log-pressure altitude, as with the Eulerian mean. More details of the relationships between these circulations are given in Section 6.5.2 of WMO (1985).

We now briefly mention the specification of two-dimensional (2D) models for transport and chemical studies. (A comprehensive account is to be found in Chapter 12 of WMO, 1985.) These tend to use either (a) the Eulerian-mean formulation or (b) the residual, diabatic or similar formulation. In each case some form of eddy flux convergence term generally appears in the tracer continuity equation. (An exception is the generalised Lagrangian-mean formulation, in which, however, a mass-continuity equation of the form (3.10) does not hold.) This eddy flux must be parameterised in terms of the zonal-mean state with which the model is working; this is usually done by means of a two-dimensional 'diffusion tensor' $\underset{\sim}{K}$, relating the vector eddy flux to the mean gradient $(\partial\overline{\chi}/\partial y, \partial\overline{\chi}/\partial z)$.

A parameterisation of this type is difficult to justify rigorously except in one formulation of type (b), based on the 'effective transport velocity' introduced by Plumb (1979). This is defined by

$$\overline{v}^T \equiv \overline{v} - \rho_0^{-1} \frac{\partial}{\partial z}(\rho_0 \Xi) \ , \quad \overline{w}^T \equiv \overline{w} + \frac{\partial \Xi}{\partial y} \ , \tag{3.16}$$

where

$$\Xi \equiv \tfrac{1}{2}(\overline{\eta' w'} - \overline{\zeta' v'}) \tag{3.17}$$

and (η', ζ') are the northward and vertical eddy parcel displacements. The diffusion tensor is given by

$$\underset{\approx}{K} \equiv \frac{1}{2}\frac{\partial}{\partial t}\begin{pmatrix} \overline{\eta'^2} & \overline{\eta'\zeta'} \\ \overline{\eta'\zeta'} & \overline{\zeta'^2} \end{pmatrix} \tag{3.18}$$

in the absence of eddy sources and sinks of tracer (i.e. $S' = 0$). Given these definitions, the tracer continuity equation becomes

$$\frac{\partial \overline{\chi}}{\partial t} + \overline{v}^T\frac{\partial \overline{\chi}}{\partial y} + \overline{w}^T\frac{\partial \overline{\chi}}{\partial z} = \rho_0^{-1}\nabla\cdot(\rho_0\underset{\approx}{K}\cdot\nabla\overline{\chi}) + \overline{S}\ , \tag{3.19}$$

with errors that are cubic in eddy amplitude, and the mass-continuity equation is analogous to (3.10). In this formulation the rate of change of $\overline{\chi}$ thus depends on advection by the effective transport velocity, diffusion by $\underset{\approx}{K}$, and the mean sources and sinks. Note that the diffusion is generally anisotropic and may even take the form of 'anti-diffusion' if, for example, all parcel displacements are decreasing with time. If S' is nonzero but weak a similar formulation applies, with additional terms appearing in $\underset{\approx}{K}$.

Owing to the Lagrangian nature of Ξ and $\underset{\approx}{K}$, it is in general difficult to estimate them from atmospheric observations. However, Plumb and Mahlman (1987) derived $(\overline{v}^T, \overline{w}^T)$ and $\underset{\approx}{K}$, month-by-month, from a three-dimensional general circulation model of the troposphere and lower stratosphere. They then used these quantities in a 2D model incorporating (3.19), which was found to reproduce quite well the zonal-mean evolution of the 3D model's tracer fields. (The errors incurred by neglecting the terms cubic in amplitude were thus mostly fairly small in this case.) Further discussion of this approach is given in Section 6.5 of WMO (1985), where it is noted, among other things, that $(\overline{v}^T, \overline{w}^T)$ and $\underset{\approx}{K}$ must be specified in advance for the 2D model, and cannot self-consistently be allowed to respond to changes in the 2D model fields: a full 3D model would be needed to calculate the required changes in the eddy statistics.

A related approach that has also been used in 2D transport models is the isentropic formulation, in which θ is used as a vertical coordinate. Under quasi-adiabatic conditions the 'vertical' velocity and displacement are small. This method has been employed by Tung (1982, 1986) and Ko et al. (1985).

Advantages of 2D models over 1D models include the facts that they can in principle represent meridional transport in a fairly self-consistent manner and can be validated against zonal-mean observations at different latitudes. Furthermore, their dynamical formulation is still sufficiently simple to allow a large proportion of the computing resources to be devoted to detailed chemistry. On the other hand they do suffer from the essentially 'non-interactive' nature of the eddy parameterisations mentioned above, which can be fully circumvented only by use of 3D models.

3.3 Three-dimensional models

We conclude this section with a few brief remarks on 3D models of middle atmosphere transport. In principle, these may treat the coupling between fully three-dimensional dynamics, radiation and photochemistry. A major advantage of 3D models is that they explicitly describe the dynamics of the large-scale planetary waves and the transport associated with them. Of course, such models still need to parameterise the effects of disturbances of smaller scale than the grid size, including gravity waves. 3D models can also represent comparatively localised events and handle the transport of local injections of tracer. This may make for more meaningful validation of the models against observational data. Unlike the 2D or 1D models, no artificial separation is made into zonal or global means and departures therefrom.

Owing to the large computer resources required for dealing with the dynamics and radiation, few 3D models have yet included any but the simplest chemical schemes. An approach that is often adopted is to perform the tracer studies 'off-line': the model's velocity fields are stored, and are used in a separate calculation to advect the tracer around. This allows dynamically-passive tracers to be treated, with or without active chemistry. However, dynamically-active tracers, which influence the motion or temperature fields, cannot be studied in this way, but must be incorporated in a fully interactive 'on-line' calculation. Examples of the off-line method include those of Mahlman and his co-workers [see, e.g., Mahlman and Moxim (1978), Mahlman, Levy and Moxim (1980, 1986), Mahlman (1985) and Plumb and Mahlman (1987)], and those of Grose and his colleagues [Grose, Turner and Nealy (1985); also this volume]. An example of the on-line approach, including a simple but accurate parameterisation of ozone photochemistry, is that of Cariolle and Déqué (1986).

Finally, it should not be forgotten that large 3D models generate vast quantities of data, which must be suitably organised and compressed if meaningful comparisons with observations are to be made, and physical interpretation carried out. There is a clear need for further discriminating diagnostics to aid this process.

4. CONCLUSION

This brief tutorial review has attempted to give a flavour of the fundamentals of atmospheric transport phenomena, and some idea of the types of model that are currently being used to study problems of chemical transport in the stratosphere and mesosphere. Other papers in this volume will consider specific applications in more detail.

REFERENCES

Andrews, D.G. and McIntyre, M.E., 1976: Planetary waves in horizontal and vertical shear: the generalized Eliassen-Palm relation and the mean zonal acceleration. J. Atmos. Sci. 33, 2031-2048.

Andrews, D.G. and McIntyre, M.E., 1978a: Generalized Eliassen-Palm and Charney-Drazin theorems for waves on axisymmetric mean flows in compressible atmospheres. J. Atmos. Sci. 35, 175-185.

Andrews, D.G. and McIntyre, M.E., 1978b: An exact theory for nonlinear waves on a Lagrangian-mean flow. J. Fluid Mech. 89, 609-646.

Brewer, A.W., 1949: Evidence for a world circulation provided by the measurements of helium and water vapour distribution in the stratosphere. Quart. J. Roy. Meteor. Soc. 75, 351-363.

Butchart, N. and Remsberg, E.E., 1986: The area of the stratospheric polar vortex as a diagnostic for tracer transport on an isentropic surface. J. Atmos. Sci. 43, 1319-1339.

Cariolle, D. and Déqué, M., 1986: Southern hemisphere medium-scale waves and total ozone disturbances in a spectral general circulation model. J. Geophys. Res. 91, 10,825-10,846.

Clough, S.A., Grahame, N.S. and O'Neill, A., 1985: Potential vorticity in the stratosphere derived using data from satellites. Quart. J. Roy. Meteor. Soc. 111, 335-358.

Dobson, G.M.B., 1956: Origin and distribution of the polyatomic molecules in the atmosphere. Proc. Roy. Soc. Lond. A 236, 187-193.

Dunkerton, T.J., 1978: On the mean meridional mass motions of the stratosphere and mesosphere. J. Atmos. Sci. 35, 2325-2333.

Garcia, R.R. and Solomon, S., 1985: The effect of breaking gravity waves on the dynamics and chemical composition of the mesosphere and lower thermosphere. J. Geophys. Res. 90, 3850-3868.

Grose, W.L., Turner, R.E. and Nealy, J.E., 1985: Transport processes in the stratosphere: model simulations and comparisons with satellite observations. Middle Atmosphere Program: Handbook for MAP, vol. 18, pp. 381-385.

Holton, J.R., 1986: A dynamically based transport parameterization for one-dimensional photochemical models of the stratosphere. J. Geophys. Res. 91, 2681-2686.

Hoskins, B.J., McIntyre, M.E. and Robertson, A.W., 1985: On the use and significance of isentropic potential vorticity maps. Quart. J. Roy. Meteor. Soc. 111, 877-946.

Jones, R.L., 1984: Satellite measurements of atmospheric composition: three years' observations of CH_4 and N_2O. Adv. Space Res. 4, No.4, 121-130.

Ko, M.K.W., Tung, K.K., Weinstein, D.K. and Sze, N.D., 1985: A zonal-mean model of stratospheric tracer transport in isentropic coordinates: numerical simulations for nitrous oxide and nitric acid. J. Geophys. Res. 90, 2313-2329.

Leovy, C.B., Sun, C.-R., Hitchman, M.H., Remsberg, E.E., Russell, J.M., Gordley, L.L., Gille, J.C. and Lyjak, L.V., 1985: Transport of ozone in the middle atmosphere: evidence for planetary wave breaking. J. Atmos. Sci. 42, 230-244.

McIntyre, M.E. and Palmer, T.N., 1983: Breaking planetary waves in the stratosphere. Nature 305, 593-600.

McIntyre, M.E. and Palmer, T.N., 1984: The 'surf zone' in the stratosphere. J. Atmos. Terrest. Phys. 46, 825-849.

Mahlman, J.D., 1985: Mechanistic interpretation of stratospheric tracer transport. Adv. Geophys. 28A, 301-323.

Mahlman, J.D., Levy, H. and Moxim, W.J., 1980: Three-dimensional tracer structure and behavior as simulated in two ozone precursor experiments. J. Atmos. Sci. 37, 655-685.

Mahlman, J.D., Levy, H. and Moxim, W.J., 1986: Three-dimensional simulations of stratospheric N_2O: predictions for other trace constituents. J. Geophys. Res. 91, 2687-2707. Also corrigenda, 91, 9921.

Mahlman, J.D. and Moxim, W.J., 1978: Tracer simulation using a global general circulation model: results from a midlatitude instantaneous source experiment. J. Atmos. Sci. 35, 1340-1374.

Murgatroyd, R.J. and Singleton, F., 1961: Possible meridional circulations in the stratosphere and mesosphere. Quart. J. Roy. Meteor. Soc. 87, 125-135.

Plumb, R.A., 1979: Eddy fluxes of conserved quantities by small-amplitude waves. J. Atmos. Sci. 36, 1699-1704.

Plumb, R.A. and Mahlman, J.D., 1987: The zonally-averaged transport characteristics of the GFDL general circulation/transport model. J. Atmos. Sci., to appear.

Pyle, J.A. and Rogers, C.F., 1980: Stratospheric transport by stationary planetary waves -- the importance of chemical processes. Quart. J. Roy. Meteor. Soc. 106, 421-446.

Taylor, G.I., 1953: Dispersion of soluble matter in solvent flowing slowly through a tube. Proc. Roy. Soc. Lond. A 219, 186-203.

Tung, K.K., 1982: On the two-dimensional transport of stratospheric trace gases in isentropic coordinates. J. Atmos. Sci. 39, 2330-2355.

Tung, K.K., 1986: Nongeostrophic theory of zonally averaged circulation. Part I: Formulation. J. Atmos. Sci., to appear.

WMO, 1985: World Meteorological Organization: Global Ozone Research and Monitoring Project. Report No. 16. Atmospheric Ozone, 1985.

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 1	June 1984	A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model. C Gordon and M Bottomley
DCTN 2	Sept 1984	The parametrization of the upper ocean mixed layer in coupled ocean/atmosphere models. C Gordon and M Bottomley
DCTN 3	Sept 1984	The response of a simple thermodynamic sea-ice model to GCM forcing appropriate for perennial sea-ice. M Bottomley
DCTN 4	Sept 1984	Potential vorticity in the stratosphere derived using data from satellites. S A Clough, N S Grahame and A O'Neill
DCTN 5	Sept 1984	Simulation of the seasonal cycle of SST using globally specified climatological data to force simple ocean models. N S Grahame
DCTN 6	Sept 1984	Development of an improved GCM radiation scheme. A Slingo and R C Wilderspin
DCTN 7	Sept 1984	On the specification of surface fluxes in coupled atmosphere/ocean general circulation models. J F B Mitchell, C A Wilson and C M Price
DCTN 8	Sept 1984	A survey of possible repercussions of the Soviet river reversal programme on Arctic sea-ice. H Cattle
DCTN 9	Oct 1984	A study of a strong stratospheric warming during January 1982 T D A Fairlie
DCTN 10	Nov 1984	An intercomparison of the Met O 11 M45 model and the Met O 20 11-level model on the Cyber 205 J F Dyson
DCTN 11	Nov 1984	Simulation of observed time series of temperature and precipitation over eastern England. D N Reed
DCTN 12	Nov 1984	Forecasts and analyses of the onset of the Southwest monsoon. R Kershaw
DCTN 13	Nov 1984	On modelling the effects of CO_2 on climate. J F B Mitchell
DCTN 14	Dec 1984	Development of a revised longwave radiation scheme for an atmospheric general circulation model. A Slingo and R C Wilderspin
DCTN 15	Jan 1985	Rossby wavetrains in the stratosphere forced by localised disturbances in the troposphere. C J Marks (Oxford), A O'Neill and V D Pope

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 16	Jan 1985	Observing system experiments using the Meteorological Office's 15-level model. R A Bromley
DCTN 17	Feb 1985	Atmospheric general circulation simulation of the climate of the Mediterranean region. H Cattle
DCTN 18	Feb 1985	Diurnal variation and cloud in a general circulation model. C A Wilson and J F B Mitchell
DCTN 19	Feb 1985	The sensitivity of the Saharan region in a general circulation model. W M Cunnington and P R Rowntree
DCTN 20	Mar 1985	The impact of aircraft observations on analyses and forecasts from the operational 15-level model. B R Barwell
DCTN 21	Mar 1985	Non-linear effects on heat-induced circulation of the tropical atmosphere. A E Gill and P J Phlips
DCTN 22	Apr 1985	Simulation of cloudiness with the U.K. Meteorological Office 11-layer AGCM. A Slingo and R C Wilderspin
DCTN 23	Apr 1985	An integrated approach to the representation of cloud processes. R N B Smith
DCTN 24	May 1985	Nuclear Winter: Calculations of the shortwave radiative effects of soot aerosols. A Slingo and P Goldsmith
DCTN 25	Jul 1985	Sea ice and the Antarctic winter circulation: a numerical experiment. J F B Mitchell and T S Hills
DCTN 26	Jul 1985	A Budyko-Sellers model for study of ocean-atmosphere interaction. W J Ingram
DCTN 27	Aug 1985	The effect of a shoaling thermocline on equatorially-trapped Kelvin waves. A E Gill and B A King
DCTN 28	Sept 1985	CO_2 climate sensitivity and model dependence of results. J F B Mitchell, C A Wilson and W M Cunnington
DCTN 29	Sept 1985	Handbook of the Meteorological Office 11-layer atmospheric general circulation model. Volume 1 : Model Description Edited by A Slingo
DCTN 30	Sept 1985	Impact of land surface variations on African rainfall in general circulation models. P R Rowntree, M F Wilson and A B Sangster

DCTN 31 Nov 1985 The annual cycle of sea level in the eastern tropical Pacific.
G R Bigg and A E Gill

DCTN 32 Nov 1985 An analytic model of the heat-induced tropical circulation in the presence of a mean wind.
P J Phlips and A E Gill

DCTN 33 Nov 1985 The effect of resolution and diffusion on the simulated climate.
J F Dyson

DCTN 34 Jan 1986 Rossby adjustment over a step.
M K Davey, A E Gill, E R Johnson and P F Linden

DCTN 35 Jan 1986 Issues concerning the evaluation of effective surface roughness of heterogeneous terrain.
D J Carson

DCTN 36 Mar 1986 Climate sensitivity and past climates : evidence from numerical studies
J F B Mitchell

DCTN 37 Mar 1986 Parametrizations of land-surface processes in Meteorological Office numerical weather prediction and climate models
D J Carson

DCTN 38 Apr 1986 Modelling of land-surface processes and their influence on European Climate
D A Warrilow, A B Sangster and A Slingo

DCTN 39 Apr 1986 Simulation of climatic change due to increased CO_2
J F B Mitchell

DCTN 40 May 1986 On the interpretation of the Eliassen-Palm flux divergence.
Part I : Theory
D G Andrews

DCTN 41 June 1986 Experiments on tropical circulation with a simple moist model.
M K Davey and A E Gill

DCTN 42 Jun 1986 Parameter regimes for studying isolated eddies.
P D Killworth and M K Davey

DCTN 43 Jun 1986 A comparison of the impact of an envelope orography and of a parametrization of orographic gravity-wave drag on model simulations.
A Slingo and D W Pearson

DCTN 44 Jul 1986 The effect of a sea-surface temperature anomaly on a prediction of the onset of the south-west monsoon over India.
R Kershaw

DCTN 45 Jul 1986 An introduction to the parametrization of land-surface processes.
D J Carson

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 46 Sept 1986 Lateral boundaries and the heat-induced flow in the tropical atmosphere.
P J Phlips and A E Gill

DCTN 47 Nov 1986 Simulation of the diurnal cycle of outgoing longwave radiation with an atmospheric GCM.
A Slingo, R C Wilderspin and S J Brentnall

DCTN 48 Feb 1987 A 2 x CO_2 climate sensitivity experiment with a global climate model including a simple ocean
C A Wilson and J F B Mitchell

DCTN 49 Feb 1987 Sensitivity of the Met O 20 11-layer model to increased land surface roughness length
C C Heasman

DCTN 50 Feb 1987 The influence of atmospheric waves on the general circulation of the middle atmosphere
D G Andrews

DCTN 51 Feb 1987 Transport mechanisms in the middle atmosphere: an introductory survey
D G Andrews